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# Soil water storage and rooting depth: key factors controlling recharge on rangelands

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## Abstract:

The practice of removing woody vegetation to enhance water supply in semiarid rangelands in the United States continues to generate considerable interest, even though past research has yielded apparently contradictory results concerning its efficacy. In an attempt to elucidate the factors that determine whether and how woody vegetation removal affects water supply, we analysed the problem using a water balance approach. In our analysis, deep drainage is the water balance component associated with water supply. Because the herbaceous vegetation that replaces the woody plants generally has a shallower effective rooting depth ($R_d$), the amount of soil water potentially available for transpiration is reduced and more is available for deep drainage. The potential increase in deep drainage can be estimated from the capacity of the soil to store plant-available water ($S_c$) and may be substantial. Our case study on sagebrush rangeland documents how $R_d$, and consequently $S_c$, changed after woody vegetation at the site was removed by burning. Using depth profiles of soil water content and matric potential, we showed that the $R_d$ of the post-fire vegetation was about 140 cm, 60 cm less than that of the pre-fire vegetation, and that this resulted in a potential increase in deep drainage of about 6 cm of water—which in semiarid rangelands is substantial. Historical precipitation patterns indicate that there is nearly always sufficient net precipitation to generate the additional 6 cm of deep drainage at this site. However, in most of the area the soil depth is less than 140 cm, so that transpiration and deep drainage would be unaffected by the vegetation change and the overall water supply enhancement would be much less than 6 cm. These results indicate that the change in $S_c$ that may follow woody shrub removal is an important criterion to evaluate sites for vegetation conversion. Copyright © 2006 John Wiley & Sons, Ltd.

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## INTRODUCTION

The past 150 years have witnessed, on a global scale, the conversion of semiarid grasslands and savannas to shrublands dominated by woody vegetation. Many factors—including increased grazing pressure, reduced fire frequency, and increased atmospheric $CO_2$—contribute to this phenomenon, which is often referred to as *woody plant encroachment* (Archer, 1994; Van Auken, 2000). This dramatic shift in vegetation cover no doubt has important, though still poorly understood, implications for both biogeochemical (Hibbard *et al*., 2001) and hydrological cycling (Huxman *et al*., 2005).

Currently, there is a good deal of interest in re-establishing the historical dominance of non-woody vegetation in semiarid landscapes. A range of motivations drive this interest, such as improving biodiversity and wildlife habitat; increasing the amount and quality of livestock forage; reducing erosion, which diminishes the productivity of the land; and augmenting water supply. The last of these, augmenting water supply, is the focus of this paper.

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The results of individual studies of this issue have been mixed, but most previous research has concluded that removing woody vegetation to increase water supply in semiarid rangelands in general is impractical (Ffolliott and Thorud, 1974; Hibbert, 1983; Baker, 1984). The basic argument for the use of this technique is that evapotranspiration—primarily interception and transpiration—is higher in shrublands than in grasslands, which makes less water available for streamflow and groundwater recharge (Hibbert, 1983). Studies in juniper shrublands, indicating that interception rates are considerably higher under juniper canopies than in inter-canopy areas (Owens *et al*., 2006; Skau, 1964), support this argument. For other shrublands, however, interception by shrubs is comparable to that by herbaceous vegetation (Wilcox, 2005). Similarly, while some studies show definite increases in streamflow and recharge as a result of reduced transpiration rates following shrub removal (Hibbert, 1983; Williamson *et al*., 2005), other studies show that the presence or absence of woody plants has relatively little effect on total evapotranspiration (Carlson *et al*., 1990; Dugas and Mayeux, 1991; Weltz and Blackburn, 1995; Dugas *et al*., 1998).

These apparently conflicting results are at least partly due to the fact that in semiarid regions, the potential for increasing water supply (streamflow and/or groundwater recharge) is small per unit area and is highly dependent on soil, vegetation, and climatic conditions, which vary widely (Phillips *et al*., 2004). A better understanding of these varied conditions would enable better, more efficient land-management decisions. Given the considerable expense and effort required for vegetation conversion, gaining such an understanding is vital. To this end, a conceptual model—or framework—for ascertaining the conditions under which woody plant cover may influence streamflow has been proposed (Huxman *et al*., 2005; Wilcox *et al*., 2006). Referred to as the *shrub-streamflow framework*, it is a qualitative tool that uses physiography, climate, and soil drainage or subsurface flow characteristics to identify areas where vegetation conversion is most likely to enhance water supply.

Our work extends that concept to include more quantitative assessment criteria. In the first part of this paper we analyse, in a general way, water balance on semiarid landscapes to explicitly identify those factors that determine whether water supply will increase if woody shrubs are removed. In the second part, we apply this approach to a case study at a site in southwest Idaho, where the burning of woody vegetation altered the water balance. We then show how increases in water supply depend on spatially variable soil properties and on seasonal climatic conditions.

## THE APPROACH

We apply the one-dimensional water balance equation considered fundamental to hydrology (Rodriguez-Iturbe, 2000; Wilcox *et al*., 2003) to inter-drainage uplands where the water table is sufficiently deep so that during dry periods it provides no water for transpiration. These areas constitute by far the greater part of the landscape surface area in semiarid regions.

Although this approach has the disadvantage that, being one-dimensional, it does not directly calculate streamflow, it has three major advantages: (1) it provides a framework that explicitly incorporates all factors affecting water supply; (2) it uses storage (capacitance) parameters that are relatively easy to quantify and are amenable to functional modelling approaches suitable for large-scale management (e.g. Addiscott, 1993; Seyfried, 2003); and (3) it can easily incorporate deterministic spatial variability (e.g. Seyfried and Wilcox, 1995), which is likely to be important in most large-scale applications.

The water balance, during a specified time period, for a volume of soil bounded at the top by the atmosphere and at the bottom by an arbitrarily defined depth, can be written as follows:

$$\Delta S = \text{Net soil water input} - \text{net soil water output} = (P - E_\text{p} - E_\text{s} - F_\text{o}) - (T + D_\text{d}) \quad (1)$$

where $\Delta S$ is the change in soil water storage; $P$ is the amount of water deposited on the plant canopy or soil surface (including litter) as rain or snow; $E_\text{p}$ is the evaporation (or sublimation) of water from the plant surface (i.e. interception); $E_\text{s}$ is the evaporation from the soil surface; $F_\text{o}$ is the net overland flow; $T$ is transpiration;



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and $D_d$ is the deep drainage—water that passes through the root zone and is assumed to eventually become groundwater recharge and/or streamflow. Units for all terms are $cm^3$ of water per $cm^2$ of soil surface.

*Deep drainage and precipitation distribution*

Although either $D_d$ or $F_o$ may generate streamflow and therefore be associated with water supply, the two result from very different processes. When past studies arrived at contradictory conclusions, part of the reason might have been a failure to distinguish between streamflow generated by $D_d$ and that generated by $F_o$. In general, $F_o$ occurs in semiarid regions when precipitation rates exceed soil infiltrability, and is not desirable because it tends to increase erosion on site and sedimentation downstream. In addition, $F_o$ is easily modified by transient 'side effects' of treatment (e.g. vehicle traffic), and any changes may be unrelated to vegetation removal *per se*. In contrast, $D_d$ occurs when net water inputs exceed the soil storage capacity. From a hydrologic perspective, it is generally the goal of semiarid vegetation management because increases in $D_d$ enhance local spring flow and streamflow and may have larger-scale effects on groundwater. We therefore regard $D_d$ as the component of the water balance that represents water supply, and on that basis we rearrange Equation (1) to isolate $D_d$:

$$D_d = (P - E_p - E_s - F_o - T - \Delta S) \tag{2}$$

Note that increases in $F_o$ may actually be associated with a reduction in $D_d$.

Equation (2) also highlights the importance of precipitation distribution in generating $D_d$. In order for $D_d$ to be greater than 0, there must be some time period when water inputs exceed the 'loss' terms plus $\Delta S$. During the warm season, $E_p$, $E_s$ and $T$ are at their maxima; $F_o$, which usually results from summer thunderstorms, tends to be greatest; and $\Delta S$ is potentially large because soils tend to be relatively dry. In contrast, during the cool season, $E_p$ tends to be much reduced as the leaf area index (LAI) declines; $E_s$ is lower owing to reduced evaporative demand; $T$ may approach zero as plants are dormant; $F_o$ is relatively rare; and soils tend to be wetter, reducing the potential contribution of $\Delta S$. Therefore, while past research (e.g. Bosch and Hewlett, 1982; Hibbert, 1983) has tended to emphasize mean annual precipitation as a criterion for evaluating a site's potential to generate $D_d$, the seasonal distribution of precipitation is at least as important.

*Soil water storage*

The $\Delta S$ term is important because it defines how much net input is required to generate $D_d$. This factor was generally overlooked in previous research, probably because it depends on antecedent conditions and is therefore difficult to quantify. However, in semiarid regions it is possible to estimate the effect of $\Delta S$ on $D_d$ in a way that incorporates the influence of vegetation change. It has been widely observed in the field that, on an annual basis, $\Delta S$ is approximately zero—that is, each year the soil water content consistently returns to the same minimum value (Campbell and Harris, 1977; Schwinning *et al*., 2005; Seyfried *et al*., 2005). This observation is consistent with the fact that, in semiarid regions, the potential evapotranspiration greatly exceeds $P$ on an annual basis. Thus, virtually all plant-available soil water is returned to the atmosphere. For there to be generation of $D_d$, therefore, there must be some period during the year in which cumulative net water inputs at least exceed the capacity of the soil to store plant-available water ($S_c$). During this time period, $\Delta S \geq S_c$.

The value of $S_c$ depends on the soil porosity ($P_s$), the effective plant rooting depth ($R_d$), and the plant-available water holding capacity of the fine-earth soil fraction ($A_w$). These are related such that

$$S_c = P_s \times R_d \times A_w \tag{3}$$

where $P_s$ is expressed as a volumetric fraction (dimensionless), $R_d$ is in centimetres, and $A_w$ is expressed as a volumetric fraction of the porosity (dimensionless). Porosity is directly related to bulk density, which tends to be related to texture, but in rangeland soils, coarse fragment content often is much more important. In the extreme case where the root zone penetrates fractured rock, the porosity may be only 0·01 to 0·13



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(e.g. Bockgard and Niemi, 2004; Zhang *et al*., 2004). Thus, although roots typically penetrate fractured rock, the effect on $S_c$ and the water balance is relatively small.

Traditionally, $A_w$ is the difference between the water content at field capacity and that at permanent wilting point (a matric potential $[h] =$ about $-1500$ kPa in agronomic settings). In rangelands, the concept also applies, although rangeland plants generally do not wilt. We use instead the term 'plant extraction limit', which may be considerably lower than $-1500$ kPa (e.g. Pockman and Sperry, 2000).

From a hydrologic perspective, $R_d$ is the maximum depth to which plants extract water to $h$ values lower than would result from drainage. In very dry environments, where no $D_d$ is generated, these $h$ values may be observed below the rooting zone (Walvoord *et al*., 2002). In these environments, fluxes are too small to be estimated by means of water balance (Seyfried *et al*., 2005). But if $D_d$ occurs, even rarely, $h$ below the rooting zone will not be much lower than about $-50$ kPa (Walvoord *et al*., 2002). Although it is not clear what root densities are required for effective use of the plant-available water, there are indications that these may be very low (e.g. Sturges, 1980).

The general observation is that, in semiarid environments, $R_d$ of woody shrubs is greater than that of grasses (Schenk and Jackson, 2002). For this reason, a reduction in $R_d$ is expected when herbaceous, grassy vegetation replaces woody vegetation. The reduction in $R_d$ leads in turn to a reduction in $S_c$ (Equation (3)) and to a lowering of the input threshold required to generate $D_d$. This is important because the change in $R_d$ resulting from removal of woody shrubs has a potentially large impact on $D_d$. For example, if $R_d$ changes from 150 cm to 75 cm following woody vegetation removal, the potential increase in $D_d$ is 10 cm of water in medium-textured soils.

The water balance analysis approach, thus, points to the importance of $S_c$ as a criterion for estimating the effect of woody shrub removal on water supply in semiarid landscapes. This is consistent with previous research that emphasized the change in $T$, which results from vegetation change (e.g. Hibbert, 1983; Turner, 1991). If all other terms in Equation (2) are unaffected by vegetation change, the reduction in $T$ is equal to the change in $S_c$. The advantage of emphasizing $S_c$ is that it provides a basis for quantifying the reduction in $T$. In some environments, notably those dominated by juniper, $E_p$ may also change substantially—in which case the effect of shrub removal will be greater than would be estimated from the change in $S_c$ alone.

In the following section we present data from a field experiment to illustrate how woody vegetation removal affected plant uptake and soil water storage and how, quantitatively, that logically impacts $D_d$. We then examine these changes in the context of climatic and landscape variability at the site.

## CASE STUDY: PRESCRIBED FIRE AT REYNOLDS CREEK, IDAHO

### *Background*

The study site is in the Reynolds Creek Experimental Watershed, located in southwest Idaho, USA. The mean annual precipitation at the site is about 550 mm; the months of July and August are typically very dry (Hanson, 2001). Vegetation at the site was dominated by a dense stand of mountain big sagebrush (*Artemesia tridentata* ssp. *vaseyana*) and bitterbrush (*Purshia tridentata*), with mountain snowberry (*Symphoricarpos oreophilus*) and Western Juniper (*Juniperus occidentalis* ssp. *occidentalis*) also present. Shrub heights ranged from 120 to 190 cm. The grass understory was dominated by bluebunch wheatgrass (*Agropyron spicatum*) and Sandberg bluegrass (*Poa secunda*).

A prescribed fire, administered by the Bureau of Land Management, was set on 17 September 2002, and allowed to burn 0·67 km$^2$. No seeding was done after the fire. The main objectives were to determine the extent to which water use by the post-fire vegetation differed from that by the pre-fire vegetation, and whether the change in vegetation would result in a water supply increase.



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*Field methods*

Before the fire, we isolated a fire exclosure about 20 m by 40 m by creating 2-m fire breaks on three sides (a dirt road served as a fire break on the fourth side). Personnel were be stationed at the site to ensure that the fire did not 'jump' these breaks. There was no discernible difference between the vegetation inside the fire exclosure and that of adjacent areas. The complex terrain and small size of the fire exclosure precluded direct measurement of evapotranspiration with methods such as eddy covariance.

Our instrumentation was designed to compare soil water dynamics under the post-fire and pre-fire (unburned) vegetation. We collected the following data in both types of vegetation: soil water content ($\theta$, $cm^3$ $cm^{-3}$) with depth, soil temperature with depth, soil water matric potential ($h$, kPa) with depth, LAI (by plant type), and snow depth. In addition, standard meteorological data were collected at a weather station located about 300 m from the site.

*Vegetation.* LAI and plant type (i.e. shrub, grass, tree, forb) were measured with a point frame (Clark and Seyfried, 2001) in twelve 1-$m^2$ plots, six in the burned area and six in the fire exclosure. Measurements were made on four dates during the growing season of the year following the fire and once, near peak growth, each subsequent year. Before installing the soil water measuring equipment, we dug two pits on each side of the fire line and about 7 m away from it. These pits, measuring roughly 1 m wide $\times$ 2 m long $\times$ 1·3 m deep, are perpendicular to the fire line and about 7 m apart. In each pit, we measured root densities in duplicate 1-l samples taken at various depths down to 140 cm. Roots were separated from the soil by rinsing with a 1-mm mesh sieve, drying, and weighing at 60 °C. Roots were visible at all depths, but densities declined dramatically with depth—from 0·025 g $cm^{-3}$ at 10 cm to 0·0025 g $cm^{-3}$ at 20 cm to less than 0·001 g $cm^{-3}$ at depths greater than 60 cm.

*Soils.* From each of the four pits described above, samples were taken for analysis of soil bulk density, and structure, texture, and colour were described. Although the pits were all somewhat different in terms of the thickness of soil horizons, they exhibited similar, distinctive horizonation. The A horizon, about 33 cm thick with a loam texture, is underlain by transitional horizons having sandy-loam to sandy-clay-loam textures to a depth of about 72 cm. These horizons are underlain by a high-clay-content ($>$35%) argillic horizon, usually to a depth of 135 cm. In two of the pits, gravel content was higher, and granitic bedrock was apparent near the pit bottom.

*Soil water content.* The soil profiles at opposite ends of each pit were instrumented with two-rod, 30-cm long time domain reflectometer (TDR) probes (eight TDR profiles in all—four in the burned area and four in the fire exclosure). The probes were inserted horizontally at depths of 10, 30, 60, 90, and 120 cm. A final probe was inserted at a 45° angle from 135 to 150 cm. The TDR probes were connected to a Campbell Scientific TDR100[1] system. TDR data were collected hourly starting 8 May 2001. The Ledieu calibration equation (Ledieu *et al.*, 1986) was used to calculate volumetric soil water content from the measured apparent dielectric permittivity.

Soil water content was also monitored by a neutron probe. Six aluminium access tubes were installed, three in the burned area and three within the fire exclosure. The maximum measurement depth was 195 cm at four sites (two in the burned area and two in the fire exclosure). The maximum measurement depth at the remaining two sites was 105 cm because rocks prevented augering deeper. Data were collected periodically during the growing season (starting 4 June 2001) at depths of 15, 30, 60, 90, 105, 120, 150, 180, and 195 cm. The instrument used was a Troxler 4300 series neutron probe. Its calibration is described by Seyfried *et al.* (2001b).

*Soil matric potential.* At the same six sites instrumented with neutron probes, tensiometers were installed at depths of 90, 120, 150, and 180 cm (only 90 and 120 cm at the two shallow soil sites), about 1 m from the neutron-probe access tubes. A Soil Measurement Systems Inc. tensimeter, which is accurate to about 0·2 kPa, was used to measure $h$. (Mention of manufacturer is only for the convenience of the reader and implies no endorsement by the author or by USDA.) Only relatively deep measurements were made because the length of time between the end of freezing temperatures and the occurrence of $h$ values below the tensiometer limits (about $-750$ kPa) is very short close to the surface. Measurements were made periodically over the entire area before the fire, and in both burned and unburned areas the year following the fire.

## RESULTS

### *Vegetation*

Essentially all vegetation in the burned area was reduced to ash by the fire. The soil surface was covered with black ash and, in some places near larger shrubs, white ash. It remained in this state through winter. By early spring, a partial cover of perennial grasses and forbs, burned but not killed by the fire, was evident. Around the larger shrubs there was a 'ring', about 30 cm in diameter, which was completely devoid of vegetation. There is some evidence of sprouting from some of the bitterbrush shrubs.

The LAI in the fire exclosure was two to three times greater than in the burned area the year following the fire (Figure 1). As the growing season progressed, LAI declined precipitously in both the burned and unburned areas, as typically happens during the hot, very dry summers in the region (Seyfried, 2003). In fall and winter, the LAI of the unburned vegetation maintains a minimum of about 0·2.

The vegetation of the burned area was dominated by forbs, which accounted for about 75% of the total LAI throughout the growing season (Table I). In the fire exclosure, the shrub component increased from 29 to 96% during the dry summer season, as the grasses and forbs died or senesced. Grasses in the fire exclosure were usually 2–3 times more abundant than forbs.

### *Soil water content dynamics*

We illustrate the internal soil water dynamics of the site with the TDR profile data from one pit in the fire exclosure during water year (WY) 2003 (Figure 2). The data gap in April was due to failure of the data

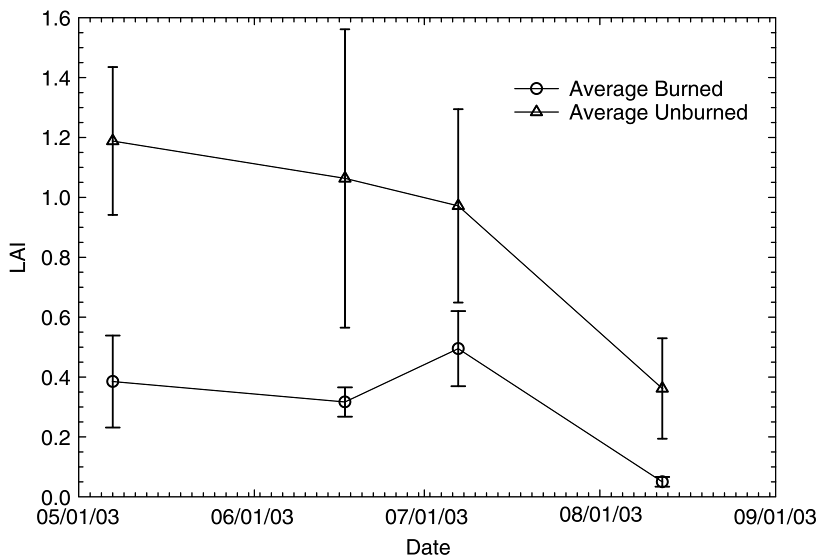


Figure 1. Leaf area index, as an average of six point-frame plots in burned area and six point-frame plots in the fire exclosure, during the growing season following the fire. Minimum (wintertime) LAI of the unburned vegetation is about 0·2



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Table I. Plant type composition as percentage of total LAI

| Date | Treatment | Composition (%) | | |
|---|---|---|---|---|
| | | Forb | Grass | Shrub |
| 7 May | Burned | 77·5 | 22·5 | 0 |
| | Unburned | 27 | 44 | 29 |
| 17 June | Burned | 66 | 34 | 0 |
| | Unburned | 14 | 32 | 54 |
| 7 July | Burned | 77 | 23 | 0 |
| | Unburned | 10 | 32 | 58 |
| 12 August | Burned | 87 | 13 | 0 |
| | Unburned | 4 | 1 | 96 |

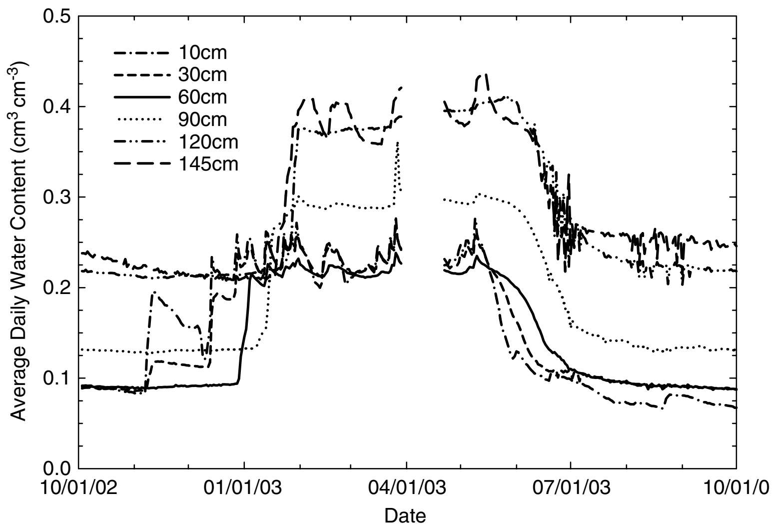


Figure 2. Example of TDR-measured soil water content from one profile in the fire exclosure for the water year 2003. Data from the deepest TDR probes, inserted at a 45° angle from 135 to 150 cm, represent conditions at 145 cm. (Note the large gap in April of 2003 data when the data logger failed)

logger. The soil water content measured during this period reflects the horizonation described above. TDR probes at the 10-, 30-, and 60-cm depths were in horizons characterized by loam textures, wherein the $\theta$ values ranged from about 0·08 $cm^3\ cm^{-3}$ to 0·25 $cm^3\ cm^{-3}$. Soil water contents measured at 120 cm and 145 cm were from a clay-loam-texture horizon and ranged from about 0·20 $cm^3\ cm^{-3}$ to 0·45 $cm^3\ cm^{-3}$. The 90-cm-depth rods were in a transitional horizon and had intermediate values.

In general, the soil water content results are consistent with the water balance approach in that: (1) the wetting front progressed vertically downward as field capacity was exceeded, and (2) $\theta$ values at the end of each WY following the fire were similar to those before the fire (i.e. $\Delta S = 0$ for the year). The lack of $\theta$ change in August and September was also observed the following year. This stability of $\theta$ within the root zone during a period of high potential evapotranspiration indicates that transpiration was limited by soil water availability (i.e. the plant extraction limit was reached). The large jump in $\theta$ at all depths (observed in both years following the fire) in response to winter rains and snowmelt indicates that input water penetrated to at least 145 cm. The neutron-probe data are consistent with these observations to depths of 195 cm, although the temporal detail in the data is lacking.



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*Soil matric potential*

Matric potentials measured during the growing season before the fire (2002), in the area that would later be designated a fire exclosure, declined rapidly and, by mid July, could not be read reliably at any depth because of cavitation. Similar $h$ dynamics, although not measured, probably occurred in the burned area that year, because the soil water dynamics were so similar. For 2003, $h$ in the fire exclosure followed the same pattern as the year before: on 10 May, it was near 0 at all depths, indicating very wet conditions (Figure 2), with a downward vertical head gradient through the profile generating $D_d$ (Figure 3). By mid July, $h$ in the fire exclosure dropped below −50 kPa at all depths (except 180 cm, which lagged slightly), indicating that plants were actively drawing water from all depths measured. In the burned area, matric potentials measured in October 2003 at 90 and 120 cm lagged slightly behind those measured at the same depths in the fire exclosure: at 150 cm they had fallen to only −35 kPa, and at 180 cm they were −10 kPa. These results indicate that water from the 90- and 120-cm depths was taken up by transpiration, whereas water from the 150- and 180-cm depths was drained by gravity. (No data were collected during the WY 2004 because many of the tensiometers were destroyed by cattle.)

*Soil water storage*

*TDR data.* The calculation of total soil water storage was based on the assumption that the water content measured by an individual TDR rod at a specific depth in the soil profile was representative of water content 10–15 cm above and below that depth. Thus, the value measured at 10 cm represented the 0- to 20-cm depth interval, that measured at 30 cm represented the 20- to 45-cm depth interval, etc. Because the soil horizons of the various profiles differ in thickness (proportions of clay- and sandy-loam textures), the total amount of water stored at any given time varied considerably among profiles. However, the variability of changes in soil water storage, a parameter critical for water balance calculations, is much less because the ranges of measured $\theta$ for the different textures are similar (Figure 2). We calculated those changes relative to the minimum value measured during the year of the fire, taken to be 0, which approximates a condition of no plant-available water.

As Figure 4 shows, before the fire the average amount of soil water stored, measured by TDR to a depth of 145 cm ($SR_{145}$) in four profiles in the areas designated for burning and four in the fire exclosure, were

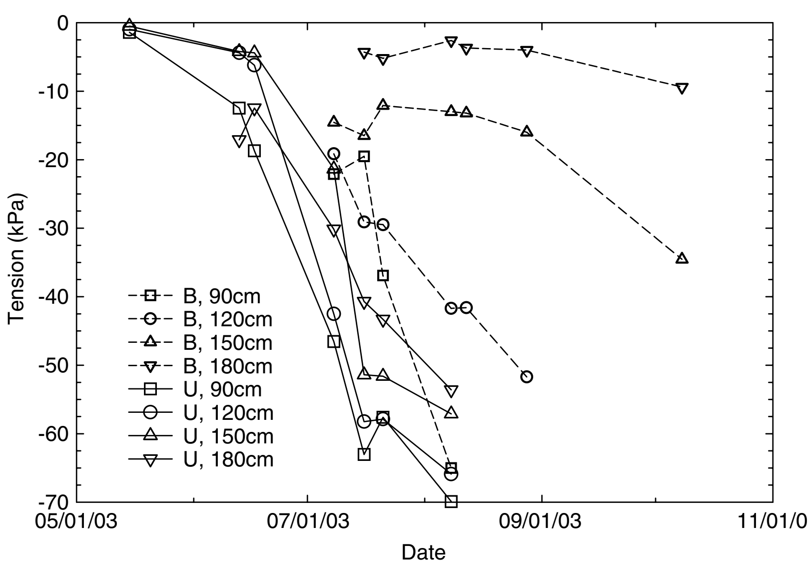


Figure 3. Average matric potential (kPa) measured during 2003 for the burned (B) and unburned (U) areas at depths of 90, 120, 150, and 180 cm



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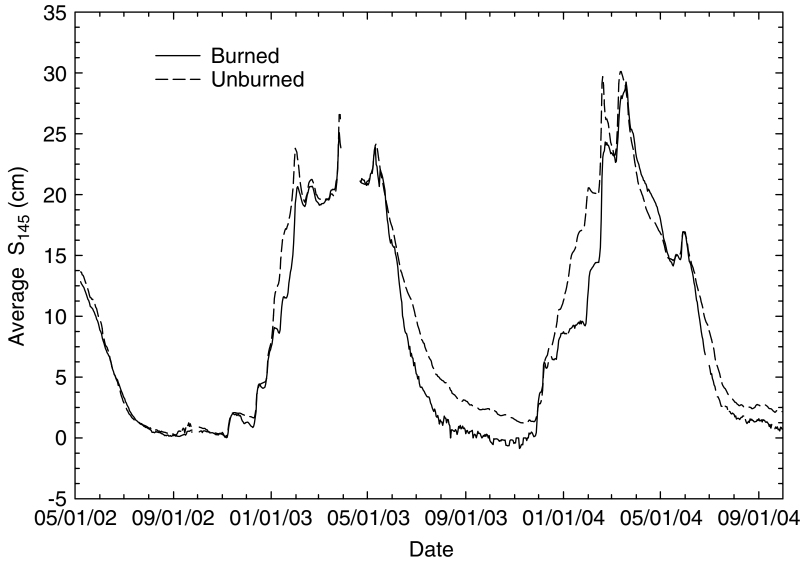


Figure 4. Soil water storage to a depth of 145 cm ($SR_{145}$) measured with TDR. The curve before 17 September 2002 (the date of the fire) represents TDR measurements from the fire exclosure area and from the area designated for burning. The latter part of the curve represents TDR measurements made after the fire. Both curves are averages—one from the four soil profiles in burned area and the other from the four profiles in the unburned area (exclosure)

essentially identical. For the two years following the fire, the overall dynamics were very similar, with peaks in late winter or early spring and declines to low values in late summer and early fall. In 2004, measurements from the eight profiles showed that soils in the burned area gained and lost water 10 to 20 days ahead of those in the fire exclosure. The winter of 2004 saw a considerable accumulation of snow (about 70 cm), which melted earlier from the more exposed, burned area—as confirmed by soil temperatures, which increase rapidly after the insulating effect of snow is removed. In contrast, during the winter of 2003 when very little snow accumulated, the changes in $SR_{145}$ in the burned area were more similar to those in the fire exclosure.

Maximum $SR_{145}$ values before the fire were similar to those measured in the burned and unburned areas after the fire, indicating that soil hydraulic properties were not affected by the fire. The minimum values in the fire exclosure returned to the same value each year, whereas following the fire those in the burned area were slightly higher, indicating that less water was being extracted by the replacement vegetation. Comparison of the minimum $\theta$ measured at each depth, pre- and post-fire, indicates that $\theta$ returned to about the same value at depths above 120 cm (Figure 5). Below 120 cm there was an increase in minimum $\theta$, and at 145 cm there was a statistically significant ($\alpha = 0{\cdot}2$) increase of about 0·05 cm$^3$ cm$^{-3}$—indicating that at that depth, less water was extracted by the vegetation after the fire than before it.

*Neutron-probe data.* The neutron-probe data are, by nature, periodic, and were not collected during the winter, so that they cannot portray soil water dynamics at the level of detail observed from TDR. The data did yield an overall $\theta$ pattern with time and depth consistent with that shown by the TDR measurements. The primary advantage of the neutron-probe data is that they are from a deeper soil profile. We calculated the SR for the neutron-probe data in the same way as for the TDR data, with the maximum depth of 210 cm related to the 15-cm measurement radius of the neutron probe and the maximum measurement depth of 195 cm.

Like the TDR results, the pre-fire neutron-probe SR measurements ($SR_{210}$) from the area designated for burning were essentially identical to those from the fire exclosure (Figure 6). The minimum measured $SR_{210}$ in the exclosure returned to practically the same value in the two following years. In the burned area, however, the minimum $SR_{210}$ values increased by about 6 cm during both years following the fire, indicating that less



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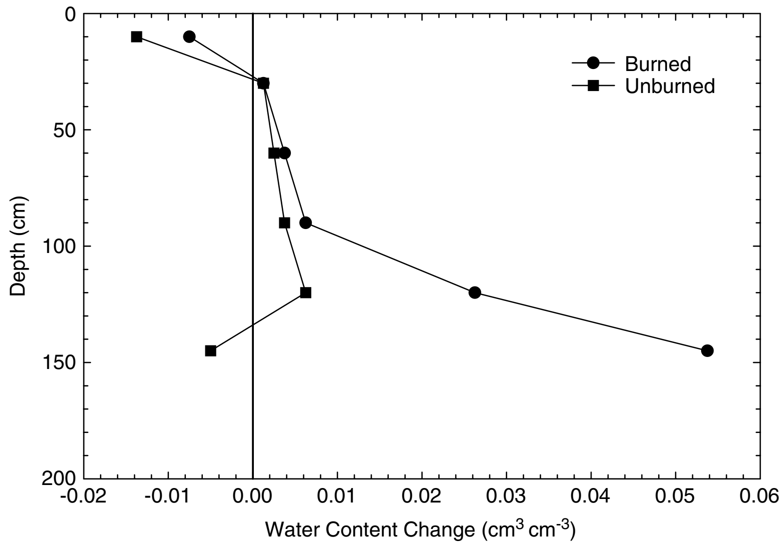


Figure 5. Average difference in water content between the minimum value measured by TDR before the fire and minimum values measured after the fire. Values of 0·01 $cm^3$ $cm^{-3}$ or less indicate that minimum values are essentially the same at that particular depth. Minimum values from the unburned fire exclosure (average of two subsequent years) did not change substantially after the fire, except possibly at the 10-cm depth, which is particularly sensitive to occasional summer rains. Minimum burned area values at 120 cm and 135 cm increased 0·025 $cm^3$ $cm^{-3}$ and 0·055 $cm^3$ $cm^{-3}$, respectively, in the 2 years following the fire

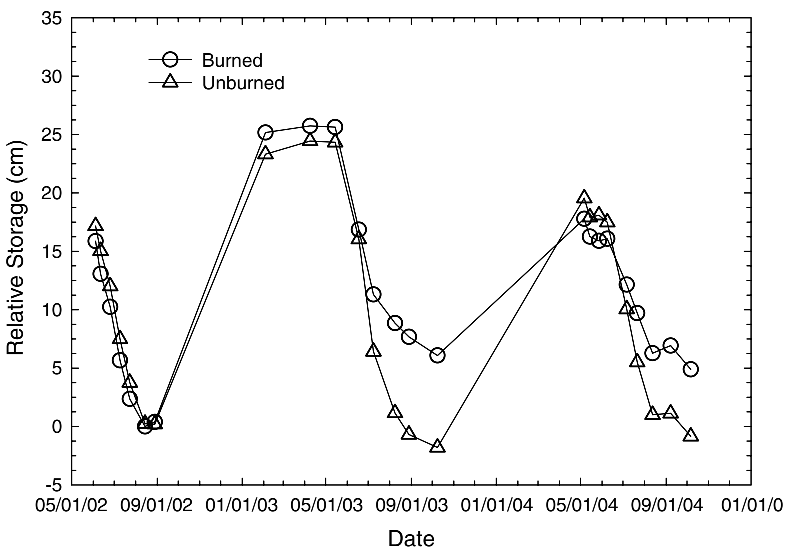


Figure 6. Average $SR_{210}$ calculated from neutron-probe measurements before and after the fire. Differences between burned and unburned areas were significant ($\alpha = 0{\cdot}2$) on the driest dates after the fire. The higher storage in burned area after the fire indicates that less water was transpired, and that given sufficient net inputs, there will be more $D_d$ the following year

water was transpired by the replacement vegetation. Differences between the plots were significant ($\alpha = 0{\cdot}2$) on the driest dates of both years.

Differences between the burned and unburned areas are primarily attributable to differences in the amount of water extracted from relatively deep in the soil profile. Minimum water contents measured during the two



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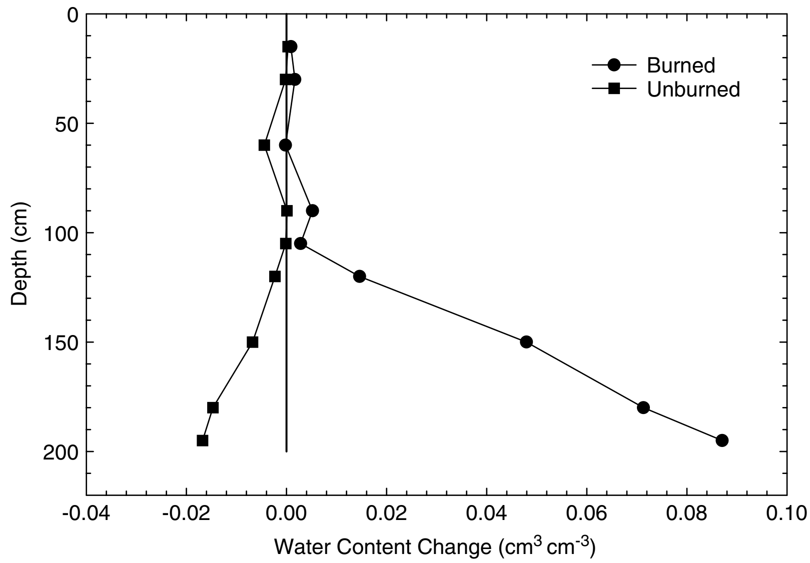


Figure 7. Average difference in water content between the minimum value measured by neutron probe before the fire and minimum values measured after the fire. Post-fire water uptake from depths above 120 cm are practically identical in the burned area to that in the unburned (fire exclosure) area. In the burned area, soil water content at depths below 120 cm was higher after the fire, indicating a decrease in water uptake from those depths

years following the fire were practically identical to pre-fire minima for all depths down to 105 cm (Figure 7). At greater depths, the minimum declined slightly in the exclosure (probably because the sampling dates were later), whereas in the burned area, it increased by greater amounts as the depth increased. The differences were significant ($\alpha = 0{\cdot}2$) at the 150-, 180-, and 195-cm depths. Thus, the measurements from all the profiles showed that water use within the first 105–120 cm of soil was the same under both types of vegetation cover and that at depths below 120 cm, less water was extracted by the replacement vegetation.

## DISCUSSION

A weakness of the experimental design employed in this study is that the soil water dynamics analysis was based on samples from only one burned and one unburned area—leaving open the possibility that observed differences were due to unknown local spatial variability and not specifically to the post-fire vegetation change. We were aware of this problem from the outset, but the logistical difficulty of controlling the fire forced us to adopt this approach. Variability within each area was taken into account by taking multiple measurements with different instruments. In addition, we believe it reasonable to attribute the observed differences to the change in vegetation for the following reasons: (1) Before the fire, vegetation cover in the areas designated for burning was indistinguishable from that in the fire exclosure, indicating uniformity of climatic and edaphic conditions; (2) Soil water dynamics in the two areas were also indistinguishable before the fire, indicating uniformity of $T$ and net $P$; (3) The observed soil water differences coincided with the abrupt and dramatic vegetation change that followed the fire.

### *Effective rooting depth*

TDR, neutron probe, and tensiometer data all indicate that the $R_d$ of the post-burn vegetation, which is dominated by forbs and grasses, was shallower than the $R_d$ of the pre-burn, woody-shrub-dominated vegetation. It is difficult to assign a specific $R_d$ to the post-burn vegetation because the boundary does not appear to be



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abrupt, or at least varies with location. There are indications of differences at 120 cm, which become more pronounced with increasing depth. The data indicate that, over the profile, there is about 12·5 cm of plant-available water per meter of soil. If $R_d$ is 140 cm, the calculated $S_c$ (Equation (3)) is 17·5 cm—which compares with an $S_c$ under unburned vegetation of 25 cm. The difference of 7·5 cm is close to what was calculated from the neutron-probe data. At these sites, $R_d$ will at some points be greater than 200 cm, but more often will be less. The LAI data indicate that by mid August, when soil water content is stable, the shrub LAI is greatly diminished, indicating drought stress; it is not likely, then, that the plants utilized deep water sources at that time.

These findings are consistent with those from other studies in semiarid locations, which explored changes in effective rooting depth following a vegetation shift. Sturges (1980), for example, working in a snow-dominated, mountain big sagebrush community, concluded that an increase in streamflow will '...only occur on land where soils are deeper than 90 cm and soil water recharge exceeds that required to fully wet the soil mantle'. Link *et al.* (1990) found that when Wyoming big sagebrush was burned and replaced with perennial grass, soil moisture below a depth of 125 cm was substantially higher than under pre-fire conditions. Williamson *et al.* (2005) found that in California, plant-available water was present below 100 cm under grass vegetation but not under native chaparral vegetation; they concluded that recharge to groundwater may occur in average to wetter years under grass vegetation. Similarly, lysimeter studies have shown significantly higher recharge from winter precipitation or snowmelt, when shrubs are replaced by annual grasses (Gee *et al.*, 1994).

### *Site evaluation*

Data on changes in effective rooting depth are much more useful when linked to $S_c$, because together they yield an estimate of the potential change in $T$—and thus in $D_d$—that results from a shift in vegetation cover. This potential change (about 6 cm of additional water at our study site) will translate to increased $D_d$ if (1) there is sufficient net water input to 'fill' $S_c$, and (2) the soils are sufficiently deep that following the shift, the new vegetation exploits a lesser soil volume than did the vegetation it replaced.

The average precipitation measured at the Reynolds Creek site, based on 10 years of record, is 54·7 cm. Of those 10 years, WY 2002 and 2003 were the two driest, with 45·7 cm and 46·8 cm, respectively. As mentioned previously, cool-season precipitation much more effectively generates $D_d$ than warm-season precipitation. This is apparent from our soil water data, which show that late spring and summer precipitation events are practically invisible because the incoming water returns to the atmosphere very quickly. In WY 2002 and WY 2003, cool-season (1 Oct–31 March) precipitation was 36·0 cm and 31·9 cm, respectively. Since water movement to the 200-cm depth was apparent in both the years, it appears that 31·9 cm of cool-season precipitation provides at least 25 cm of net water input at the site. Of the 10 years of record, only one had less cool-season precipitation (29·3 cm). The probability is high, then, that the first condition for generating $D_d$—sufficient net water input—will be met at the site almost every year.

Fulfillment of the second condition, related to soil depth, is much less likely. According to a soil map of the region (Seyfried *et al.*, 2001a), soils deeper than 150 cm are found in only about 15% of the burned area; most of the soils are less than 100 cm deep. For this reason, we expect that, for most of the burned area, the change in vegetation type will bring about no change in $D_d$. The estimated increase of 6 cm would apply to only a very small portion of the area.

We should note that, in most cases, $R_d$ will tend to increase with time (that is, the woody vegetation will return in the absence of additional management). Ultimately, the rate of this return is critical because it determines the frequency of required treatments. At present at our site, it appears that $R_d$ of the post-fire vegetation will remain about the same for at least several years. Numerous sagebrush seedlings are currently visible in the area and many bitterbrush shrubs were not killed in the fire and are sprouting, but the shrub contribution to the overall cover is very small. We will continue to monitor the site to determine what the return interval is.

*Streamflow/groundwater recharge*

Although we did not monitor streamflow or groundwater recharge in our study, data from other studies are consistent with our overall findings concerning the relationships between vegetation cover change and $D_d$ and water. For example, it has been demonstrated that in Mediterranean shrublands in the United States (California), Australia, South Africa, and Spain (Hibbert, 1983; Puigdefabregas and Mendizabal, 1998; van Wilgen *et al*., 1996; Zhang *et al*., 2001), streamflow is relatively sensitive to vegetation conversion. In contrast, in regions where warm-season precipitation constitutes a higher proportion of the annual total—such as the piñon-juniper shrublands of the southwest USA—the response of streamflow to removal of woody vegetation has been modest to none (Collins and Myrick, 1966; Myrick, 1971; Clary *et al*., 1974). In addition, when streamflow does occur in these landscapes, it tends to be generated mainly as Horton overland flow (Wilcox *et al*., 2003) or saturation overland flow (Lopes and Ffolliott, 1993), rather than from deep drainage. Similarly, in many mesquite rangelands deep drainage is insensitive to changes in woody plant cover (Weltz and Blackburn, 1995).

## SUMMARY AND CONCLUSIONS

Semiarid rangelands are, by definition, water-limited environments. Reducing woody vegetation cover can enhance water supply, but only under specific conditions. Our analysis of the issue, based on the one-dimensional water balance equation, is intended to identify those conditions and provide a means of quantitatively estimating the impact of woody vegetation removal on water supply.

We established $D_d$ as the component within the water balance framework to be associated with water supply. In order to generate $D_d$ in semiarid regions, there must be some period within the year when net precipitation inputs (i.e. total precipitation—evaporative losses) exceed $S_c$. For a given amount of precipitation, the net input is much greater in the cool season because losses are lower so that $D_d$ is more likely in climates dominated by cool-season precipitation. Because woody vegetation typically has deeper roots than the vegetation that replaces it, $S_c$ is reduced with the new vegetation, potentially increasing $D_d$ substantially. This $D_d$ increase is matched with a reduction in $T$ and can be estimated from soil ($A_w$) and plant ($R_d$) characteristics. On the basis of this analysis, the amount of $D_d$ increase in response to woody vegetation removal will be equal to the change in $S_c$ if there is a period during the year in which the net precipitation inputs exceed the $S_c$ of the new vegetation and if the soil depth is greater than the $R_d$ of the new vegetation.

A case study from southwest Idaho was used to demonstrate how a reduction in woody plant cover might affect $R_d$, $S_c$, and water supply. Data on soil water contents and matric potentials were collected in adjacent areas of the study site one year before and two years after woody vegetation was removed from one of the areas by prescribed fire. These data were used to demonstrate how $S_c$ and $R_d$ might influence $D_d$. The $R_d$, as indicated by soil water storage and $h$ dynamics, changed from about 200 cm before the fire to about 140 cm after the fire. The result was a decrease in $S_c$ of about 6 cm—implying a corresponding increase in $D_d$ during years when precipitation is sufficient for the wetting front to reach 200 cm. Climatic data from the site indicate that wetting-front penetration to 200 cm will occur at the site virtually every year. Precipitation inputs at the site are greatest during the cool season, when evaporative demand is low, and thus are particularly effective at generating $D_d$. However, because soil depths within most of the burned area are shallower than the $R_d$ of the post-fire vegetation, the overall average $D_d$ is expected to be much less than 6 cm.

These results suggest that the $S_c$ change following woody shrub removal is an important—though often overlooked—criterion for evaluating whether vegetation conversion will enhance water supply, because: (1) a reduction in $R_d$ is to be expected in most cases; (2) this reduction may result in a relatively large potential increase in water supply; and (3) the actual (as opposed to potential) effect is of conversion on water supply is sensitive to soil, plant, and climate characteristics, providing a means of discriminating among different sites.



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